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CHANNEL ADJUSTMENT OF AN UNSTABLE COARSE-GRAINED STREAM: OPPOSING TRENDS OF BOUNDARY AND CRITICAL SHEAR STRESS, AND THE APPLICABILITY OF EXTREMAL HYPOTHESES

ANDREW SIMON* AND COLIN R. THORNE

Department of Geography, University of Nottingham, Nottingham, NG7 2RD, U.K.

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ABSTRACT

Channel adjustments in the North Fork Toutle River and the Toutle River main stem were initiated by deposition of a 2·5 km^3 debris avalanche and associated lahars that accompanied the catastrophic eruption of Mount St. Helens, Washington on 18 May 1980. Channel widening was the dominant process. In combination, adjustments caused average boundary shear stress to decrease non-linearly with time and critical shear stress to increase non-linearly with time. At the discharge that is equalled or exceeded 1 per cent of the time, these trends converged by 1991–1992 so that excess shear stress approached minimum values.

Extremal hypotheses, such as minimization of unit stream power and minimization of the rate of energy dissipation (minimum stream power), are shown to be applicable to dynamic adjustments of the Toutle River system. Maximization of the Darcy–Weisbach friction factor did not occur, but increases in relative bed roughness, caused by the concomitant reduction in hydraulic depths and bed-material coarsening, were documented.

Predictions of stable channel geometries using the minimum stream power approach were unsuccessful when compared to the 1991–1992 geometries and bed-material characteristics measured in the field. It is concluded that the predictions are not applicable because the study reaches are not truly stable and cannot become so until a new floodplain has been formed by renewed channel incision, retreat of stream-side hummocks, and establishment of riparian vegetation to limit the destabilizing effects of large floods. Further, prediction of energy slope (and consequently stream power) by the sediment transport equations is inaccurate because of the inability of the equations to account for significant contributions of finer grained (sand and gravel) bank materials (relative to the coarsened channel bed) from bank retreat and from upstream terrace erosion.

KEY WORDS extremal hypotheses; Mt. St. Helens, Washington; Toutle River; sediment transport; river channel adjustments; stream power; flow resistance; hydraulic geometry

INTRODUCTION

In alluvial channels, the adjustment of geometry and hydraulic variables was expressed by Inglis (1947) in terms of transportation of load with the minimum expenditure of energy, and by Leopold *et al.* (1964) such that the rate of work accomplished by the system is a minimum. The concept of minimum energy dissipation rate in stream channels and the closely related concepts of minimum stream power and minimum unit stream power are used to describe stable alluvial-channel geometries based on the premise that, for a given water and sediment discharge and other 'constraints', such as bed-material particle size and hydraulic roughness, an alluvial channel will establish a width, depth and gradient such that:

1. it results in a minimum stream power (Chang, 1979, 1980) or unit stream power expenditure (Yang, 1976; Yang and Song, 1979) per unit length of channel; and
2. total energy loss is minimized (Song and Yang, 1980).

* On leave from U.S. Geological Survey, Raleigh, North Carolina, U.S.A.

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These minimizations and their supporting interpretations are termed 'extremal hypotheses' (Davies and Sutherland, 1983; Griffiths, 1984; Bettess and White, 1987) and may be considered analogous to Kirkby's (1977) 'maximum sediment efficiency'. In contrast to the extremal hypotheses that rely on minimization, Davies and Sutherland (1983) recommend maximization of the Darcy–Weisbach friction factor to explain alluvial channel behaviour and stable channel geometry.

Extremal hypotheses rely on either available force or flow resistance criteria for the prediction of stable alluvial-channel geometries. Field data on the mutual adjustment of many variables are severely restricted because unless the system is evolving rapidly, the time scale for monitoring adjustment process is impractical. The eruption of Mount St. Helens, Washington, and the consequent disruption of the Toutle River system in 1980 has provided the opportunity to collect and analyse a data set that includes hydraulic, sediment transport, sedimentologic, and morphologic information. Rapid and dramatic changes in channel morphology from 1980 to 1992 make the Toutle River system a natural laboratory in which to investigate the validity of extremal hypotheses on a temporal basis using both force (or stream energy and stream power) and resistance criteria.

BACKGROUND

Using the concept of minimization of stream power (or rate of energy dissipation), Chang and Hill (1977), Chang (1979, 1980), Yang *et al.* (1981) and Thorne *et al.* (1988) estimated stable alluvial-channel geometries. Stable channel geometries observed in the field or in laboratory flumes were then used to test the extremal hypotheses. These studies can be considered time-independent and often hold one or more dependent variables constant to simplify computations and interpretations (Chang, 1980; Davies and Sutherland, 1983; Snow and Slingerland, 1987). However, in nature, alluvial channels attain equilibrium conditions through time as flows interact with the boundary sediments through mutual adjustment of all dependent variables. The channel passes from the transitional state of disequilibrium towards a new equilibrium with few, if any, of the dependent variables remaining constant through time for a given water discharge. Different variables or constraints can shift in dominance and dependency and can even become extraneous during different phases of adjustment (Simon, 1992).

Studies of self-formed alluvial channels have used non-linear decay functions that approach minimum values and minimum variance (Leopold and Langbein, 1962) with time to describe the behaviour of unstable alluvial channels. For the purpose of this study, minimization is defined in terms of a non-linear decay function which asymptotically approaches some minimum value with increasing time. Variables that have been used to demonstrate minimization include sediment discharge (Parker, 1977), stream power (Bull, 1979; Simon, in press), entropy production (Karcz, 1980), relative degradation (Begin *et al.* 1981; Williams and Wolman, 1984; Simon, 1992), channel gradient (Simon and Robbins, 1987) and Froude number (Jia, 1990). To avoid misinterpretations of process–form interactions and stable channel morphologies, and to accurately account for changes in variables as the channel approaches a stable condition, it is important to test extremal hypotheses using temporally based data from unstable channels undergoing adjustment.

Using a time-dependent approach, Simon (1992) showed that minimization of the rate of energy dissipation was a unifying characteristic of channel adjustment in two extremely different, unstable fluvial systems. Analyses provided evidence of the decrease of available flow energy with time and a physically based rationale for the dominance of various processes during channel adjustment. However, because many geomorphic processes are characterized by considering the imbalance between the magnitude of the available erosional force acting on the channel boundaries and the erosional resistance provided by those boundaries, a means of accounting for the resistance of the boundary sediments would aid our understanding of channel adjustment with time. In studies of channel adjustment, boundary shear stress and critical shear stress can be used to represent the opposing erosional and resisting forces. Channel adjustment towards dynamic equilibrium then becomes a matter of achieving a balance between the forces of erosion and erosional resistance, although extreme flow events can temporarily disrupt temporal trends towards renewed stability.

STUDY AREA

On 18 May 1980, 282 km^2 of the upper North Fork Toutle River drainage basin was devastated by the catastrophic eruption of Mount St. Helens, Washington (Figure 1). Gravel-bed streams were buried beneath a 2·5 km^3 rockslide-debris avalanche (here referred to as debris avalanche) associated with the failure of the north flank of the volcano. The depth of burial was in excess of 140 m in some areas (Figure 2) but the average was about 40 m. A virgin landscape, devoid of stream channels and composed of material ranging in size from silt to boulders, was created in the upper North Fork Toutle River valley. Lahars emanated from water-saturated parts of the debris avalanche and resulted in extensive flooding and deposition of up to 4 m of sand-sized material along downstream reaches and along the Toutle River main stem. The lower North Fork Toutle River and the Toutle River main stem were initially transformed from low sinuosity gravel- and cobble-bed streams to straighter, braided, sand-bed channels with boundaries smoothed by blankets of deposited sand. Initial channel widths ranged from about 20 to 30 m at the lower end of the debris avalanche, to about 55 m near the mouth of the North Fork Toutle River, and to about 100 m at the mouth of the Toutle River main stem.

Because of the enormous sediment loads and the potential for catastrophic breaching of several debris dams, a sediment retention structure (SRS) was constructed on the North Fork Toutle River just above the mouth of the Green River. The SRS became operational in November 1987 and has a mean-annual trap efficiency of about 90 per cent. It releases relatively sediment-free water that entrains sand and fine gravel-sized material from downstream reaches.

DEVELOPMENT OF A NEW DRAINAGE SYSTEM AND CHANNEL EVOLUTION

Channels began to form on the debris avalanche surface in June 1980 from the filling and spilling of lakes that had formed in depressions and along the margins of the deposit (Janda *et al.*, 1984). The formation of these channels represented the initial phases of channel development in the upper North Fork Toutle River. By December 1982, drainage network integration had increased the contributing drainage area of the upper North Fork Toutle River Basin from 80 km^2 to its pre-eruption value of 282 km^2.

Channel evolution followed a five-stage sequence of morphologic development (Table I) that was dominated by channel widening. This sequence and the spatial trends of aggradation and degradation are very similar to those described by Simon (1989a) for the channelized sand-bed streams of West Tennessee. Bed-level adjustment in the Toutle River system is described by a dimensionless coefficient (a) that represents the dimensionless elevation of the channel bed (z/z_0) when bed level stabilizes with time and the following exponential equation approaches its asymptote (Simon, 1992) (Figure 3):

$$z/z_0 = a + b\mathrm{e}^{(-kt)} \qquad (1)$$

where z = the elevation of the channel bed, in metres above sea level, z_0 = the initial elevation of the channel bed, in metres above sea level, and a, b and k = dimensionless coefficients determined by regression. The expression $1 - a$ further represents the dimensionless amount of elevation change during initial and/or secondary phases of vertical adjustment (Figure 3b). Secondary vertical adjustments represent the temporally and spatially damped oscillations that shift between degradation and aggradation in adjusting alluvial systems (Hey, 1979; Alexander, 1981; Janda *et al.*, 1984; Simon, 1992, in press). The relative magnitude of secondary vertical responses was on average 61 per cent less than the magnitude of the initial vertical process (standard error = 4·6 per cent). Examples of adjustments to active-channel width are shown in Figure 4.

STUDY APPROACH

This paper describes channel adjustments along selected reaches of the North Fork Toutle River and Toutle River main stem observed between 1980 and 1992 in the context of flow energy and shear stress relations. Reaches of the North Fork Toutle River marked by monuments were selected for analysis (Figure 1) on

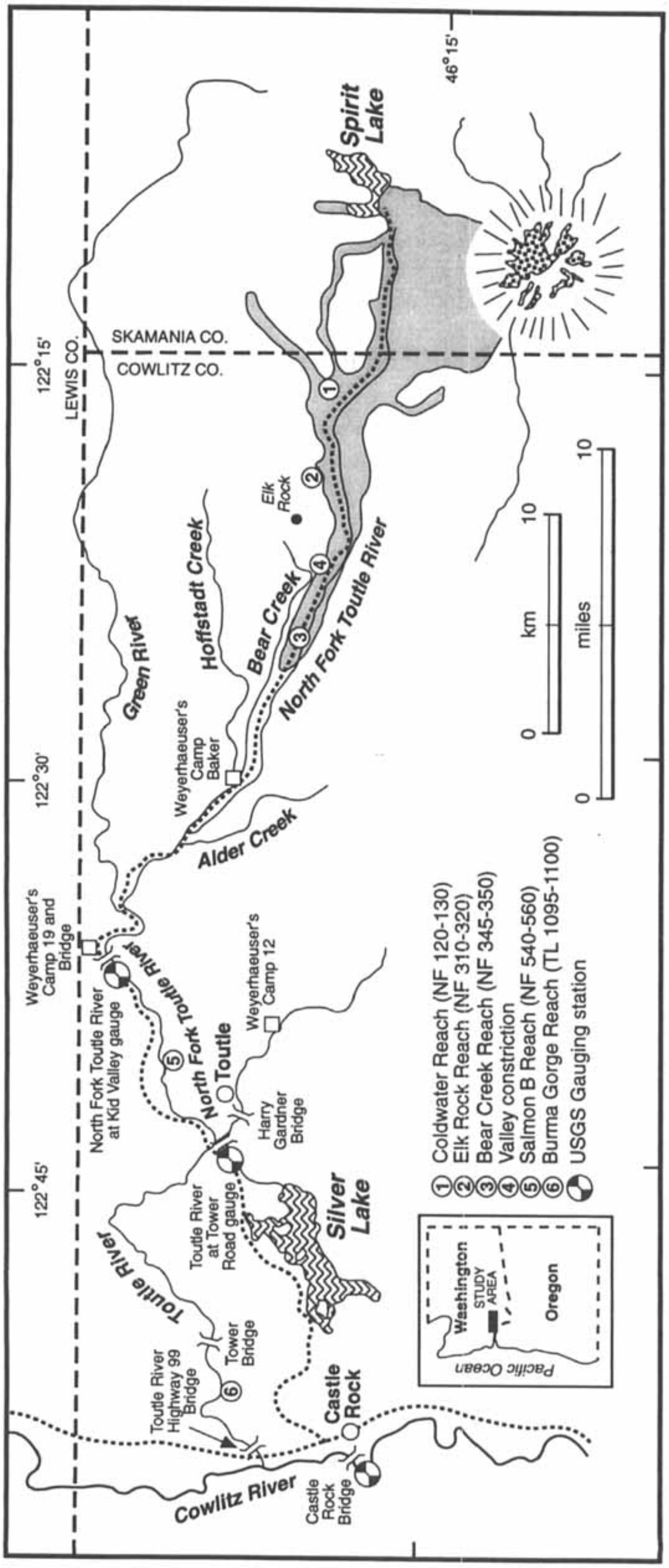


Figure 1. Location map of the Toutle River system, Washington State, USA

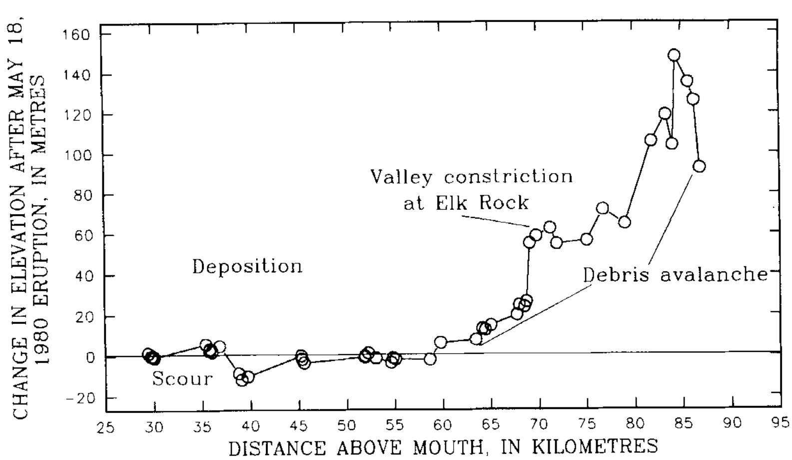


Figure 2. Elevation changes in the North Fork Toutle River valley resulting from the massive debris avalanche deposited during the eruption of Mount St. Helens on 18 May 1980. Data from surveyed cross-sections during summer and autumn 1980, and from topographic maps derived from June 1980 aerial photography

Table I. Generalized conceptual model of channel evolution for streams of the Toutle River system. Note that the model assumes an initial increase in stream-energy conditions

Stage number	Stage name	Location relative to AMD[1]	Dominant processes	Characteristics
1	Formation/ disruption	Upstream/ downstream	Scour or fill; widening or narrowing	Imposed increase in stream-energy conditions
2	Vertical I	Upstream	Degradation; net erosion	Decrease in channel gradient; flushing of silts and sands; coarsening of bed; reduction in velocity head
2	Vertical I	Downstream	Aggradation; net deposition	Decrease in hydraulic depth; increase in hydraulic roughness
3	Lateral	Upstream/ downstream	Channel widening; rate of initial vertical response wanes; net erosion	Large increase in width; reduction in hydraulic depth, increase in roughness; reduction in velocity head
4	Vertical II	Upstream	Aggradation; channel widening; net erosion	Rates of bed-level change 50–60% less than during stage 2; reduction in hydraulic depth and velocity head
4	Vertical II	Downstream	Degradation; channel widening; net erosion	ditto; all components of total-mechanical energy decrease
5	Stabilization	Upstream	Scour and fill; valley side-wall collapse and retreat; net erosion	Reduced rates of channel widening; initiation of flood-plain development
5	Stabilization	Downstream	Scour and fill; no net erosion or deposition	Reduced rates of channel widening

[1] AMD = Area of maximum disturbance, just upstream from the toe of the debris avalanche deposit

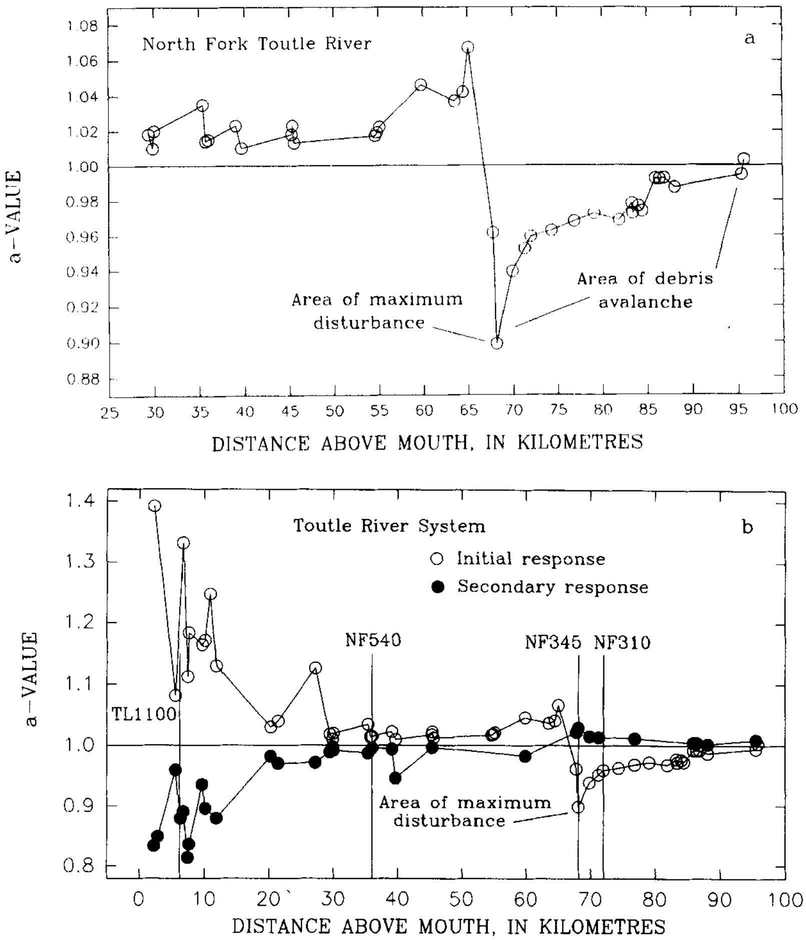


Figure 3. Bed-level response for (a) the North Fork Toutle River, and (b) the Toutle River system. Note: $a > 1{\cdot}0$ = aggradation, $a < 1{\cdot}0$ = degradation

the basis of:

1. availability of 1980 post-eruption data on channel geometry and bed-material particle size;
2. frequent surveys of the two paired cross-sections which defined the reach;
3. several measurements of bed-material particle size during phases of channel adjustment;
4. representation of aggrading and degrading conditions; and
5. the reach was straight.

Reaches are located on the debris avalanche deposit of the upper North Fork Toutle River and the lahar-affected lower North Fork Toutle River and Toutle River main stem. Significant characteristics of the selected reaches are listed in Table II.

Channel geometry

Data on 1980 post-eruption channel geometry were obtained from either field surveys or from large-scale topographic maps (1 : 4800) derived from June 1980 aerial photographs. All remaining data on channel geometry came from repeated field surveys. These surveys yielded the primary empirical data input to a one-dimensional (1-D) flow model used to estimate the hydraulic characteristics of the ungauged reaches through time.

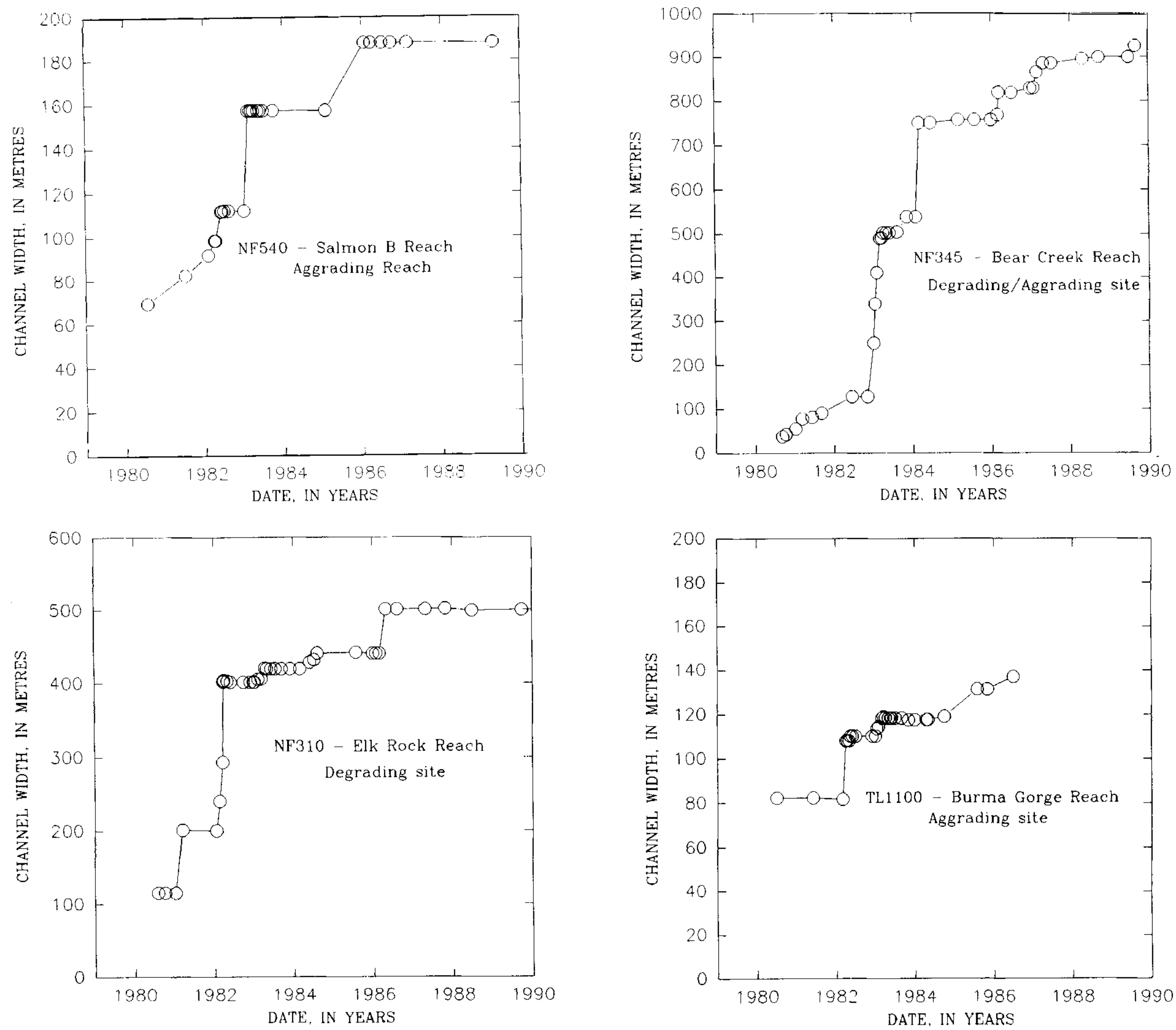


Figure 4. Examples of channel widening, Toutle River System, 1980–1992

Bed material

Sampling of surface bed material was carried out by the U.S. Geological Survey and the U.S. Army Corps of Engineers between 1980 and 1988. Efforts by the U.S. Geological Survey were concentrated at gauging stations using conventional bulk samplers, such as the BM-54 (Dinehart *et al.*, 1981; Dinehart, 1986, 1992a). Bulk sampling of bed material by the U.S. Army Corps of Engineers was conducted at various times along the North Fork Toutle River and Toutle River main stem (U.S. Army Corps of Engineers, 1988). When the streambed was dominated by gravel- and cobble-sized material, pebble-count methods were used. These data are included in Bradley *et al.* (1982) and Childers *et al.* (1988). Grid (pebble-count) sampling using frequency by number is considered equivalent to conventional bulk sieve analysis by weight for coarse mobile-pavement materials (Kellerhals and Bray, 1971).

Two samples of 50 particles each were sampled by pebble-count method in conjunction with bulk sampling of mobile-pavement materials during 1991–1992. Mobile-pavement material was sampled on low-bar surfaces along two 30·5 m transects placed parallel to a straight section of the low-flow channel. Intermediate particle diameters were measured every 0·61 m. Sub-pavement materials were sampled in

Table II. Characteristics of sites and reaches referred to in text

Site[1]	Reach name	River kilometer[2]	Disturbance type	pre-disturbed gradient	Imposed gradient	1980 D_{50} (mm)	1991–1992 D_{50}[3] (mm)	Initial vertical adjustment
TL1100	Burma Gorge	6·2	Lahars	·003	·001	0·13	56·8	Aggradation
TL1095	Burma Gorge	6·7	Lahars	·003	·001	0·13	62·7	Aggradation
NF560	Salmon B	35·4	Lahar	·007	·007	0·50	39·5	Aggradation
NF540	Salmon B	36·0	Lahar	·008	·004	0·50	44·9	Aggradation
NF350	Bear Creek	67·8	Debris avalanche, Lahar	·016	·032	0·53	32·0	Degradation
NF345	Bear Creek	68·1	Debris avalanche, Lahar	·015	·017	0·53	28·1	Degradation
NF320	Elk Rock	71·3	Debris avalanche	·022	·031	0·53	50·2	Degradation
NF310	Elk Rock	72·0	Debris avalanche	·023	·028	0·53	16·0	Degradation
NF130	Coldwater	76·8	Debris avalanche	·025	·028	0·90	25·9	Degradation
NF120	Coldwater	79·1	Debris avalanche	·028	·035	0·90	15·6	Degradation

[1] TL = Toutle River main stem; NF = North Fork Toutle River
[2] Distance measured from mouth of Toutle River main stem
[3] Data based on pebble counts of mobile-pavement materials

bulk at three locations along the transect closest to the stream (stations 0 m, 15·2 m and 30·5 m), passed through a 90 mm sieve, and dried and sieved in the laboratory. Mean weight of the samples was 228 kg. Sub-pavement materials with intermediate particle diameters greater than 90 mm were weighed and measured in the field and included as the coarsest fractions of the particle-size distribution. No size classes were truncated.

Hydraulic information

Hydraulic information (depth, shear stress, water-surface width, velocity, and energy slope) for the study reaches was estimated using WSPRO, a fixed-bed, steady 1-D step-backwater model (Shearman *et al.* 1986; Shearman, 1990). Time-series data for a given reach were generated by running the model for each time period with channel geometry data from repetitive surveys. Flow discharges representing the range of the flow duration were modelled. The Q_1 discharge (discharge that is equalled or exceeded 1 per cent of the time) was selected as a representative measure of channel-forming flows because temporal changes in hydraulic characteristics of the Q_1 discharge reflect changes in bed elevation and active-channel width. The Q_1 discharge is about five times the mean annual flow at the Kid Valley gauging station (Figure 1). Because results from the range of discharges showed parallel temporal trends, only results from the Q_1 discharge will be discussed in detail.

Estimates of Manning's n coefficient were required by WSPRO as an input parameter. These data were obtained from multiple-regression analyses of n with water-surface slope and hydraulic radius data from 537 discharge measurements between 1980 and 1990 at the Kid Valley and Tower Road gauging stations (Figure 1). The resulting equation and the r^2 value (0·57) are similar to Jarrett's (1985) equation. The regression equation used in this study is defined by (standard error = 0·00496):

$$n = 0{\cdot}143 s_{\mathrm{w}}^{0{\cdot}34} R^{0{\cdot}16} \tag{2}$$

where n = Manning's hydraulic roughness coefficient, S_{w} = water-surface slope, and R = hydraulic radius (in m).

GOVERNING EQUATIONS

The three alternative extremal hypotheses for predicting channel geometries in self-formed alluvial channels identified by Bettess and White (1987) are:

1. minimum stream power and its equivalent, minimum rate of energy dissipation;
2. minimum unit stream power; and
3. maximum friction factor.

The time rate of energy dissipation per unit channel length can be defined in terms of total stream power per unit channel length (Bagnold, 1960):

$$\Omega = \gamma w y v S = \gamma Q S \tag{3}$$

where Ω = total stream power unit length of channel (in N/s), γ = specific weight of water (in $\mathrm{N\,m^{-3}}$), w = water-surface width (in m), y = hydraulic depth (pressure head) (in m), v = mean flow velocity (in $\mathrm{m\,s^{-1}}$), Q = water discharge (in $\mathrm{m^3\,s^{-1}}$) and S = energy slope (in $\mathrm{m\,m^{-1}}$).

In this paper, head loss expressed as energy slope is used to represent the rate of energy dissipation and to test the first hypothesis by examining whether temporal changes can be expressed by non-linear decay functions that become asymptotic. Total-mechanical energy per unit weight of fluid (called total head) at a given cross-section is the sum of gravitational-potential head (elevation of bed plus depth of flow) and velocity head. If the flow is uniform:

$$H = z + y + (\alpha v^2/2g) \tag{4}$$

where H = total-mechanical energy per unit weight of fluid (total head) (in m), z = mean channel-bed elevation (water-surface elevation minus flow depth) (in m), α = coefficient for non-uniform distribution of velocity, and g = acceleration due to gravity (in $\mathrm{m\,s^{-1}\,s^{-1}}$). Head loss represents the loss (dissipation) of total-mechanical energy per unit weight of fluid between two cross-sections that define a reach:

$$h_f = [z_1 + y_1 + (\alpha_1 v_1^2/2g)] - [z_2 + y_2 + (\alpha_2 v_2^2/2g)] \tag{5}$$

where h_f = total head loss (in m). Dividing head loss by the reach length then gives energy slope or the rate of energy dissipation per unit length of channel. Since head loss is the energy dissipated at a constant value of γQ, h_f can be considered analogous to the rate of energy dissipation. Head loss was modelled for successive time periods using channel geometry data from the periodic surveys to test for minimization of the rate of energy dissipation.

To test the minimum unit stream power hypothesis, stream power per unit weight of water (ω_w; termed 'unit stream power' by Yang and Song (1979)) is calculated from:

$$\omega_w = \Omega/(\gamma y w) = vS \tag{6}$$

and plotted with time to test for non-linear decay to an asymptote.

The third extremal hypothesis identified by Bettess and White (1987) is maximization of the Darcy–Weisbach friction factor (Davies and Sutherland, 1983). For wide channels this is given by:

$$f = 8 g w^2 y^3 S/Q^2 \tag{7}$$

Equations 3, 4 and 6 represent alternative measures of the flow energy available to perform work on the channel and to entrain or transport sediment. The ability to transport sediment can also be expressed in terms of average boundary shear stress. If the flow is uniform:

$$\tau = \gamma R S \tag{8}$$

where τ = average boundary shear stress (in $\mathrm{N\,m^{-2}}$) and R = hydraulic radius (in m).

To investigate the relative role of changing bed-material characteristics on channel adjustment, the resisting force opposing boundary shear stress is expressed in terms of critical shear stress. A Shields-type criterion developed by Wiberg and Smith (1987) is used. Non-dimensional critical shear stress $(\tau_*)_{cr}$ is obtained graphically using a non-dimensional particle diameter (K_*) as the abscissa in a Shields-type entrainment

curve, where $(\tau_*)_{cr}$ is the ordinate (Wiberg and Smith, 1987):

$$K_* = 0{\cdot}0047(\zeta_*)^{1/3} \tag{9}$$

where K_* = bed roughness scale (k_s) for sediment of specific gravity 2·65 and fluid temperature of 10°C:

$$\zeta_* = \frac{D^3(\rho_s - \rho)}{\nu^2 \rho} g \tag{10}$$

where ρ = fluid density (in $\mathrm{kg\,m^{-3}}$), ρ_s = sediment density (in $\mathrm{kg\,m^{-3}}$), g = acceleration due to gravity (in $\mathrm{m\,s^{-1}\,s^{-1}}$) and ν = kinematic viscosity (in $\mathrm{m^2\,s^{-1}}$). The non-dimensional parameter (K_*) is a function only of grain size and particle density, and of fluid density and viscosity such that any grain will have a unique value of ζ_* in a particular fluid environment. The median particle size of the bed material is used as D in Equation 10.

The relative balance or imbalance between available force and resistance is found by comparing average boundary shear stress and critical shear stress. A non-dimensional excess shear stress (τ_{*e}) is computed:

$$\tau_{*e} = \tau/\tau_{cr} \tag{11}$$

where τ_{cr} = critical shear stress (in $\mathrm{N\,m^{-2}}$).

Minimum stream power approach (Chang, 1980)

In Chang's (1980) approach to the analysis of flow and sediment processes in gravel-bed streams, the equation of minimum stream power may be coupled with equations of flow resistance, sediment transport and bank slope to produce rational equations for stable channel geometry. Chang (1980) presents an account of his method, which is summarized here.

The flow resistance equation used is that of Bray (1979):

$$1/f^{0\cdot5} = 1{\cdot}36(y/D_{50})^{0\cdot281} \tag{12}$$

where f = Darcy–Weisbach friction factor.

Parker (1978, 1979) developed sediment transport equations specifically for rivers with coarse beds of mobile pavement where the bed load is calculated from:

$$\Phi = 11{\cdot}2(1/\psi - 0{\cdot}03)^{4\cdot5}\psi^3 \tag{13}$$

$$\Phi = q_s/\{g(\rho_s - \rho)/\rho]D_{50}^3\}^{0.5} \tag{14}$$

$$1/\psi = \tau/[g(\rho_s - \rho)D_{50}] = \tau_* \tag{15}$$

where Φ = bed load intensity, ψ = flow intensity (particle weight/shear force), q_s = unit bed load discharge, ρ_s = sediment density, ρ = fluid density, and τ_* = non-dimensional shear stress. Parker's (1978, 1979) equation is intended for use in the range $0{\cdot}03 > \tau_* < 0{\cdot}1$. Chang (1980) extended the range to include steeper slopes by modifying the Parker equation to produce:

$$\Phi = 6{\cdot}62[(1/\psi) - 0{\cdot}03]^5\psi^{3\cdot9} \tag{16}$$

based on Meyer-Peter's data. However, the range of prescribed sediment inflows, an important component in the minimum stream power approach, is not specified by Chang (1980).

The bank slope is also required as an input in Chang's approach. A range of bank slopes is used by Chang (1980) to characterize the general trend in gravel-bed rivers, with bank slope always below the angle of repose, but increasing as bankfull discharge increases.

The three independent equations of flow resistance, sediment transport and minimum stream power can be solved to produce regime-style relations based on empirical curve fits of numerical results for the stable channel width and depth (Chang, 1980; Thorne *et al.*, 1988):

$$w = [3{\cdot}101 + 0{\cdot}4053\ln(0{\cdot}0002383 D_{50}^{1\cdot15}/SQ^{0\cdot42})^2]Q^{0\cdot47} \tag{17}$$

$$y = [0{\cdot}2829 + 0{\cdot}0569\ln(0{\cdot}0000989 D_{50}^{1\cdot15}/SQ^{0\cdot42})]Q^{0\cdot42} \tag{18}$$

where w = channel width (in m), y = mean channel depth (in m), D = particle diameter (in m), S = channel gradient (in m m,$^{-1}$) and Q = bankfull discharge (in $m^3 s^{-1}$). The equations are intended for use in channels with mobile-bed material at bankfull flow. Chang (1980) and Thorne *et al.* (1988) noted that at steep slopes, width increases rapidly with increasing slope, which was interpreted as tending towards a braided condition. Consequently, a particular strength of his method is that it should be able to account for adjustments resulting in both single-thread and braided channel configurations.

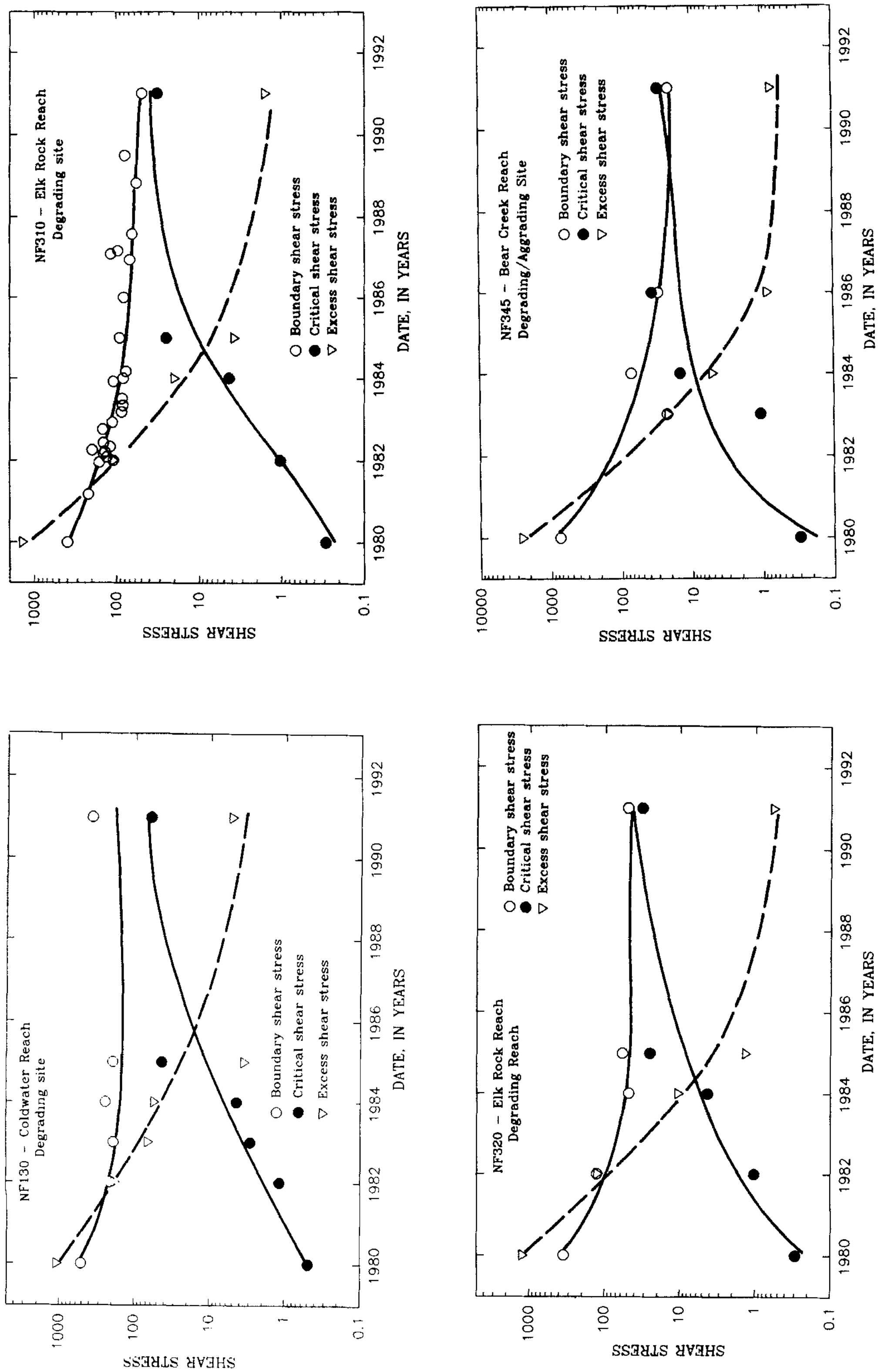
RESULTS

Boundary and critical shear stress

Average boundary shear stress decreases non-linearly and critical shear stress increases non-linearly with time for modelled reaches in the Toutle River system (Figure 5). Temporal trends of dimensionless excess shear stress (τ_{*e}) approach minimum values as defined earlier, and reflect the opposing non-linear convergence of available boundary shear and critical shear stress at the Q_1 discharge. Although there is some data scatter around the fitted curves shown in Figure 5, opposing trends of boundary and critical shear stress approach equality with time in aggrading as well as degrading reaches. These results are not indicative of channel stability because they represent only one of a range of discharges. However, they offer a physically based explanation for (1) reductions in sediment discharge with time during channel evolution in the Toutle River system (Table III) and in other disturbed fluvial systems (Parker, 1977; Simon, 1989b; Pierson *et al.*, 1992), and (2) tendencies for attaining stable bed levels as τ_{*e} approaches 1·0 with increasing time. Aggradation or degradation on the channel bed can then be interpreted in terms of whether $\tau_{*e} = 1{\cdot}0$ is overshot or not reached during different phases of adjustment.

Over the period of study, boundary shear stress at the Q_1 discharge decreased by about one order of magnitude at all but one site to tens of newtons per metre squared, while critical shear stress increased by about two orders of magnitude, also to tens of newtons per metre squared (Table IV). Increases in critical shear were the result of erosion of sand and fine gravel, leaving a coarser-grained mobile pavement that was more resistant to entrainment. This coarsening is shown by the line representing a ten-point moving average in Figure 6. Variation around the moving average can be explained by: (1) possible inconsistent sampling environment (bar or bed, for example), and (2) bed-material fining during and immediately after high flows as a result of sand deposition; subsequent low flows eroded the sand leaving the coarser, mobile pavement. To place results using the Q_1 discharge in perspective, boundary-shear-stress data for discharges ranging from $Q_{0{\cdot}5}$ to Q_{75} are shown for site NF310, Elk Rock Reach in Figure 7. Trends of boundary shear stress at all discharges are parallel and decrease non-linearly with time, from 453 to 59 N m^{-2} at the $Q_{0{\cdot}5}$, and from 88 to 27 N m^{-2} at the Q_{75}. Variations around the general trends were caused by large flow events, lake breaches, a 1982 lahar, and the pumping of water out of debris-dammed lakes. By 1991, non-dimensional excess shear stress (from Equation 11) for all discharges approached values ranging from 0·8 at Q_{75} to 1·8 at the $Q_{0{\cdot}5}$. Although these values indicate a depositional environment at Q_{75} and an erosional environment at $Q_{0{\cdot}5}$, the fact that τ_{*e} has, in all cases, decreased by two to three orders of magnitude indicates the approaching of equilibrium conditions over the entire range of the flow duration (Figure 8). These data further indicate that channel widening along the Elk Rock reach by fluvial undercutting and subsequent collapse may now be limited to the upper 25 per cent of the flow duration where values of τ_{*e} exceed 1·2, as specified by Parker (1978). In other reaches, such as Bear Creek and Salmon B, reductions in τ_{*e} at the Q_1 to values less than 1·2 by the mid- to late-1980s offers a physical explanation for reduced rates of widening in these reaches during this period (Figures 4 and 8).

The general increase in bed-material size and critical shear stress indicates that, with time, a smaller proportion of flows was capable of eroding the channel bed. Reductions in boundary shear stress occurred by rapid and preferential erosion of channel banks by fluvial undercutting and subsequent mechanical failure because (1) τ_{*e} greatly exceeded 1·2 for prolonged periods over a range of discharges, and (2) the banks became finer grained (although still non-cohesive sand and gravel) relative to the coarsening channel bed. Reductions in average boundary shear stress were, therefore, due largely to channel widening (hundreds


NF310 - Elk Rock Reach
Degrading site
Boundary shear stress
Critical shear stress
Excess shear stress
SHEAR STRESS
DATE, IN YEARS
NF345 - Bear Creek Reach
Degrading/Aggrading Site
Boundary shear stress
Critical shear stress
Excess shear stress
SHEAR STRESS
DATE, IN YEARS
NF130 - Coldwater Reach
Degrading site
Boundary shear stress
Critical shear stress
Excess shear stress
SHEAR STRESS
DATE, IN YEARS
NF320 - Elk Rock Reach
Degrading Reach
Boundary shear stress
Critical shear stress
Excess shear stress
SHEAR STRESS
DATE, IN YEARS

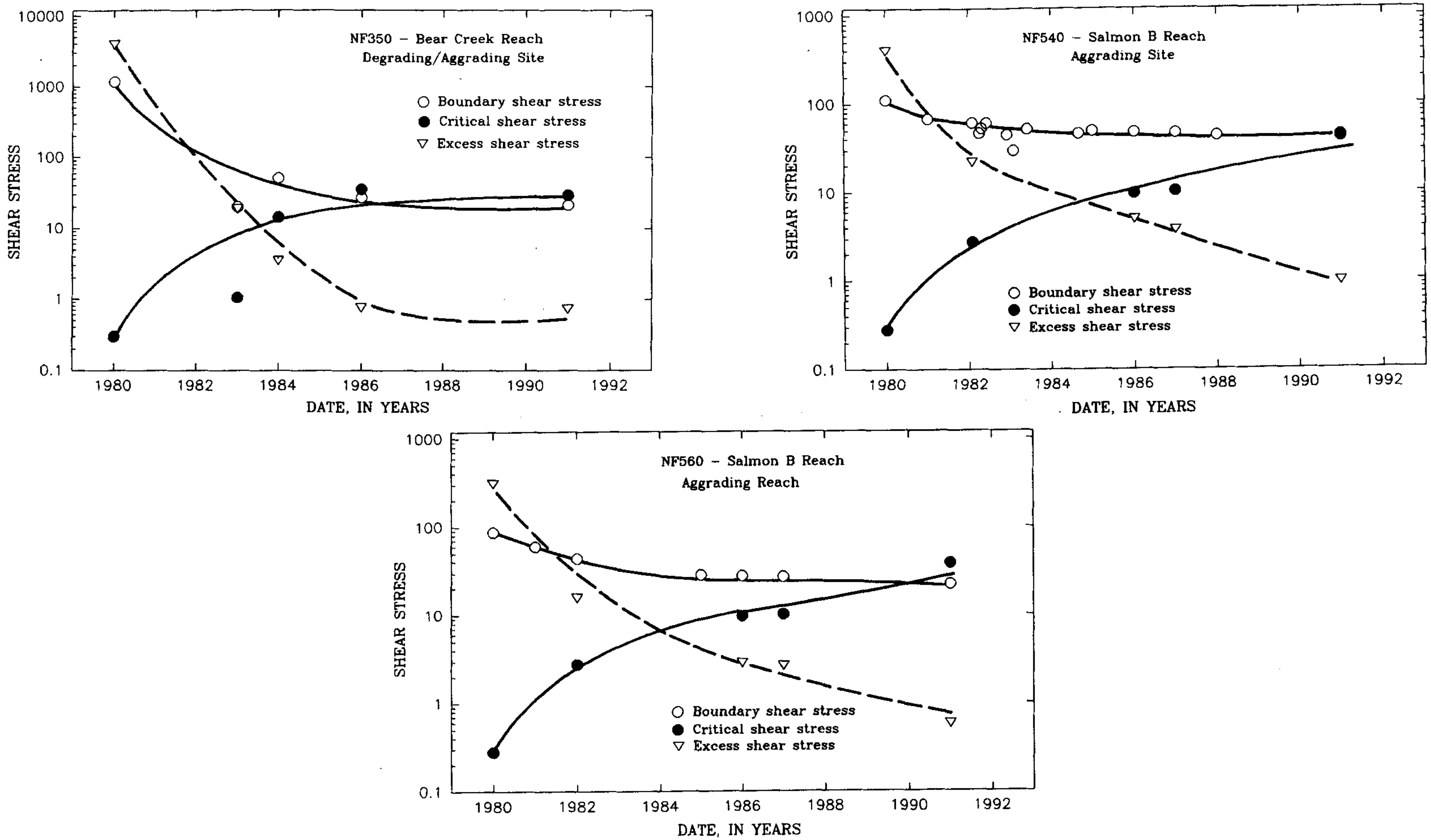


Figure 5. Trends of average boundary shear stress at the Q_1 discharge (τ), critical shear stress (τ_{cr}), and excess shear stress ($\tau_{*e} = \tau/\tau_{cr}$) for aggrading and degrading sites. τ and τ_{cr} expressed in newtons per metre squared; τ_{*e} is dimensionless. Note the opposing non-linear convergence of τ and τ_{cr}, and the minimization of τ_{*e}

Table 3. Total sediment discharge from the Toutle River system, 1980–1992. Measured data from samples taken at Toutle River at Tower Road, except where indicated. All data expressed in millions of tonnes

Year	Suspended-sediment discharge	Suspended + estimated bed-load discharge	Sediment retained by SRS[3]	Total sediment discharge
1980	136[1,2]	—	—	136[1,2]
1981	29·7[2]	33·3	—	33·3[2]
1982	40·7	45·6	—	45·6
1983	39·7	44·5	—	44·5
1984	24·7	27·7	—	27·7
1985	9·37	10·5	—	10·5
1986	7·63	8·55	—	8·55
1987	8·77	9·83	—	9·83
1988	2·20	2·47	7·31	9·78
1989	0·773	0·867	4·62	5·49
1990	2·38	2·67	9·11	11·8
1991	2·61	2·93	10·9	13·8
1992	0·743	0·833	7·70	8·53

[1] Value obtained from average of two estimates of sediment discharge on 18–19 May 1980, supplemented with measured data from samples taken at Toutle River at Highway 99
[2] Data from Toutle River at High way 99
[3] SRS = Sediment retention structure

of metres) with a consequent increase in wetted perimeter and reduction in hydraulic depth. Bed degradation (as much as 30 m) reduced channel gradient and energy slope and was also important in producing the reductions observed in average boundary shear stress. Channel widening was particularly important in reducing boundary shear stress in aggrading reaches to offset brief increases in channel gradient and energy slope.

Opposing trends of boundary and critical shear stress along the North Fork Toutle River and the Toutle River main stem indicate that bed-material coarsening with time is an integral part of channel adjustment and represents an additional degree of freedom in the Toutle River system. Increases in critical shear stress as a result of bed-material coarsening exceed reductions in average boundary shear stress by a factor of about 1·8. Results demonstrate that development of a mobile pavement on the channel bed must be accounted for in assessing expected changes in channel geometry and sediment loads in highly unstable coarse-grained alluvial streams. Gomez (1983) also came to the conclusion that temporal variations in bed-load transport rates were, in part, a function of progressive armouring on the channel bed. If coarse-grained sediments (gravel and cobbles) had not been available, it is probable that increases in critical shear

Table IV. Average boundary and critical shear stresses at the Q_5 discharge for sites in the Toutle River system, showing relative amounts of change between 1980 and 1991

Site	Year	τ (N/m^2)	Percent decrease	τ_c (N/m^2)	Percent increase	$\Delta\tau_c/\Delta\tau$[1]
NF120	1980	436	—	0·488	—	—
NF120	1991	221	49·3	62·3	127	2·58
NF310	1980	395	—	·291	—	—
NF310	1991	49·4	87·5	32·1	109	1·25
NF345	1980	747	–	·300	—	—
NF345	1991	20·7	97·2	29·1	96·0	0·99
NF540	1980	111	—	·279	—	—
NF540	1991	41·7	62·4	43·4	155	2·48
AVERAGE			74·1		122	1·83

[1] Value represents the percent change in critical shear stress divided by the percent change in boundary shear stress, from 1980 to 1991

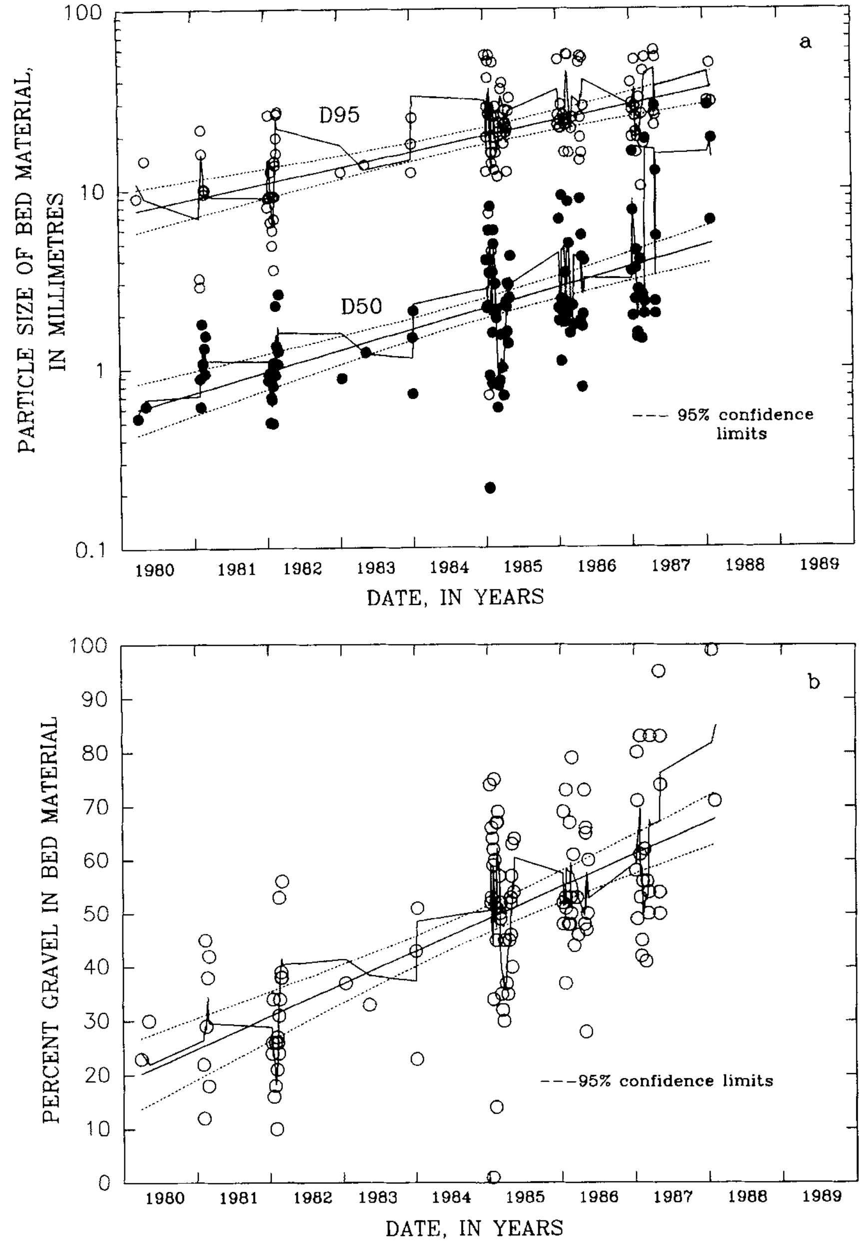


Figure 6. General coarsening of bed material at the North Fork Toutle River at Kid Valley gauging station (a) D_{95} and D_{50}; and (b) percentage gravel. Plotted line represents ten-point moving average

stress would have been limited and, therefore, geometric changes would have occurred at rates greater than those actually observed, to reduce values of boundary shear stress to levels commensurate with the erosional resistance of the finer bed sediments. Under this scenario, morphologic changes would be more drastic and cause high sediment discharges to persist for longer periods.

Minimization of unit stream power and rate of energy dissipation

Minimization of unit stream power and minimization of the rate of energy dissipation, both measures of flow energy available to perform geomorphic work, are shown to be applicable to channel adjustments in

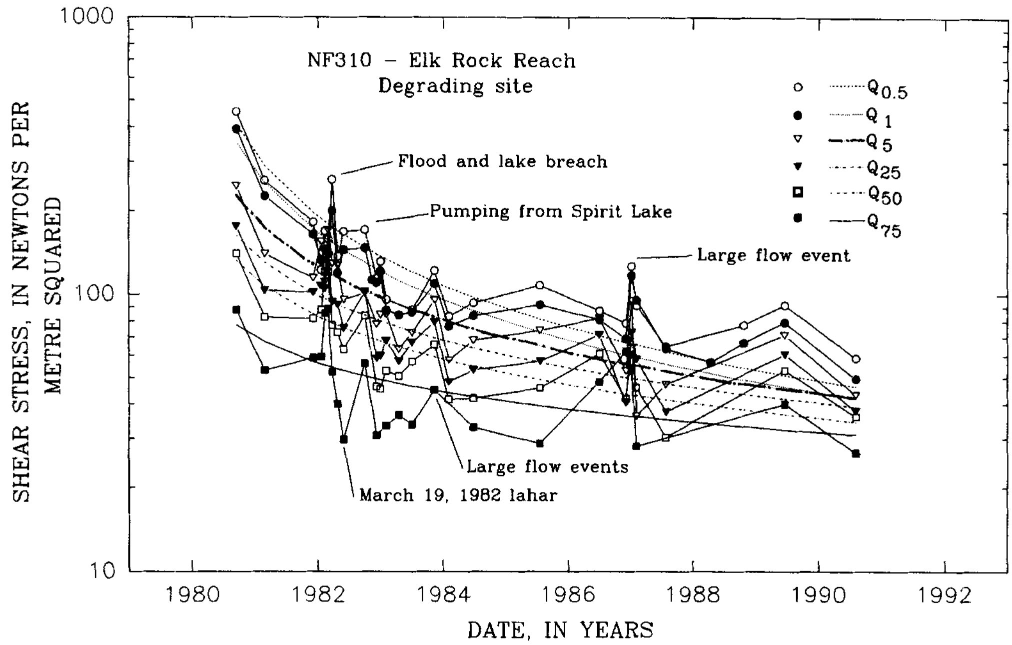


Figure 7. Average boundary shear stress for NF310, Elk Rock Reach, for a range of discharges (Q_{75} to $Q_{0\cdot05}$)

aggrading and degrading reaches of the Toutle River system (Figures 9 and 10). The minimization of unit stream power (considered as a measure of the kinetic energy component of total-mechanical energy) can provide an empirical basis for estimating reductions in sediment discharge with time (Yang, 1984). Analysis of the temporal variation of individual components of total-mechanical energy and head loss (rate of energy

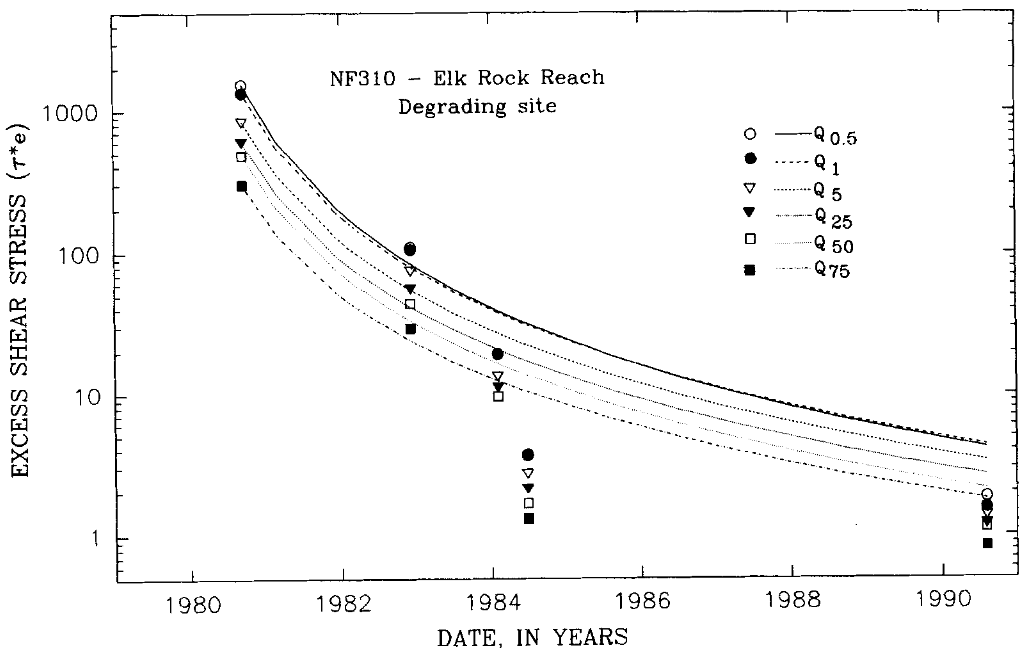


Figure 8. Excess shear stress ($\tau_{*e} = \tau/\tau_{cr}$) for NF310, Elk Rock Reach, for a range of discharges (Q_{75} to $Q_{0\cdot5}$). Note reductions to near unity

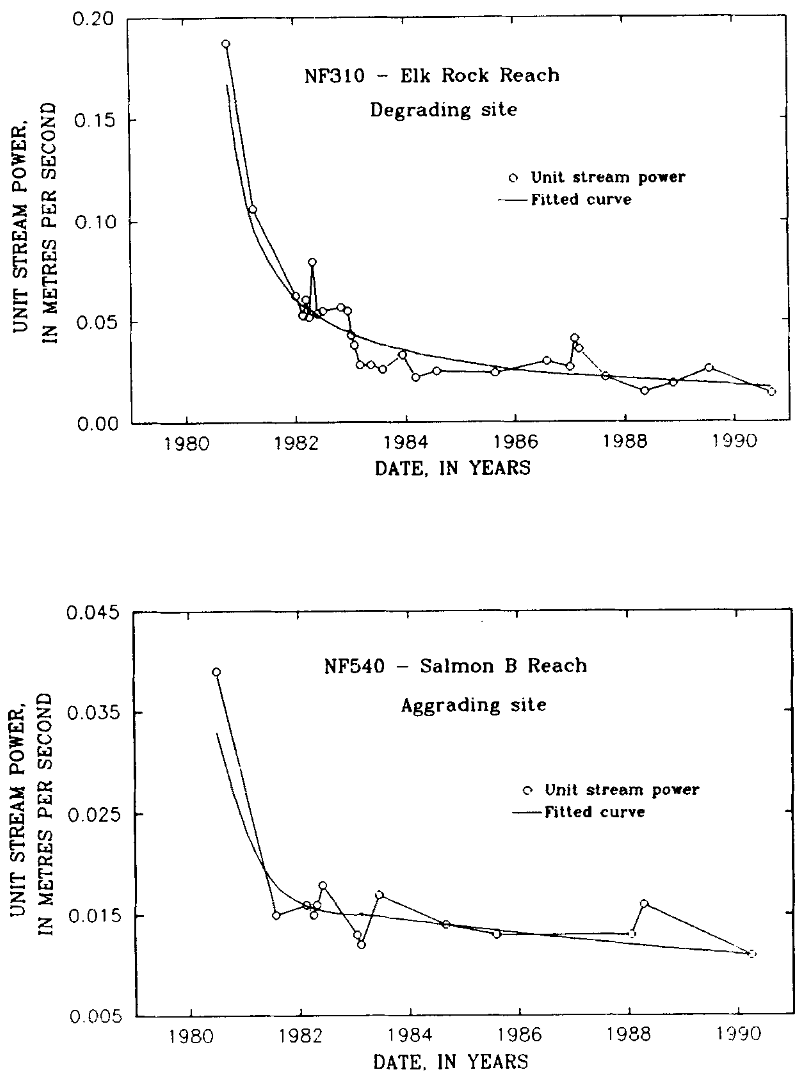


Figure 9. Trends of unit stream power at the Q_1 discharge showing minimization with time for aggrading and degrading reaches

dissipation; Equations 4 and 5) permit the identification of important processes during fluvial adjustment (Simon, 1992).

Morphologic adjustments were dominated by large increases of channel width caused by mechanical failure and fluvial entrainment of the cohesionless bank materials. For a given discharge (in this case the Q_1), increases in width cause reductions in hydraulic depth (local potential energy) and increases in relative roughness (y/D_{50}). Decreases in velocity head (kinetic energy) are a consequence of this increase in y/D_{50} as well as increases in hydraulic roughness caused by the enlargement of the wetted perimeter. On average, changes in channel width outpaced changes in bed elevation by a factor of 60 to 100 (Simon, 1992). Reductions in channel gradient appear to be of secondary importance in this fluvial environment, and are in agreement with the previous findings of Leopold and Bull (1979) and Simon (1992). It is chiefly by the process of channel widening that hydraulic depths are maintained or decreased along degrading reaches (Figure 11). Degradation without widening would lead to increases in hydraulic depth because incision does not occur equally across the channel bed but is instead concentrated at the location of maximum boundary shear and unit stream power. This holds even with concomitant decreases in velocity because the kinetic energy

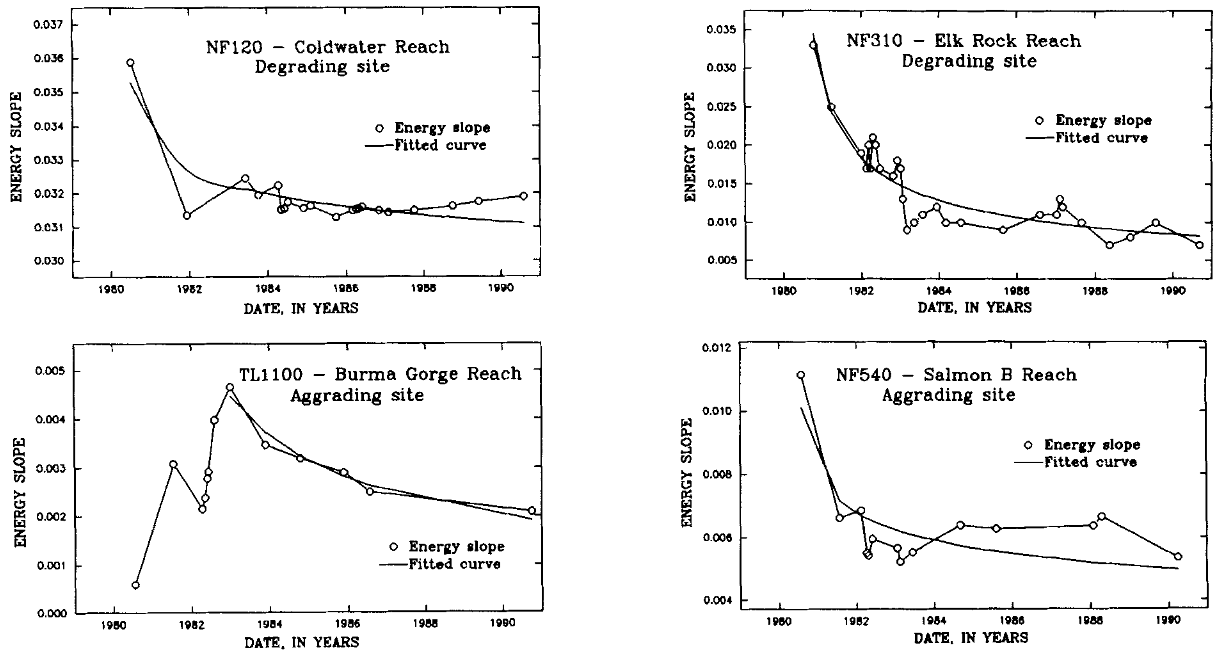


Figure 10. Minimization of total-mechanical energy (head loss) with time at the Q_1 discharge for aggrading and degrading sites

component is small relative to the hydraulic depth component (Simon, 1992). Consequently, if accompanied by channel widening, degradation appears to be the most effective means of minimizing available flow energy because all components of total-mechanical energy decrease simultaneously (Equation 4).

Channel widening also dominated aggrading reaches. Mean annual rates over the period of study reached more than 40 $\mathrm{m\,a^{-1}}$. Widening hastened reductions in hydraulic depth and counteracted the effects of potential increases in channel gradient on velocity head by producing a larger wetted perimeter with increased relative roughness. Results from the Toutle River system display the dominance of channel widening and

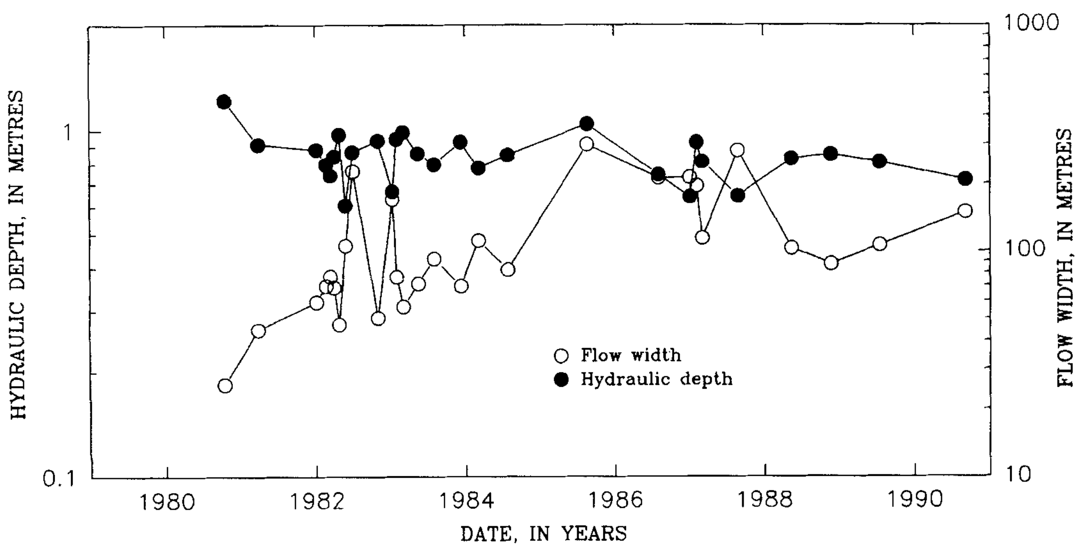


Figure 11. Temporal changes in flow width and hydraulic depth at the Q_1 discharge, NF310, Elk Rock Reach, North Fork Toutle River

support Carson's (1984) argument that width increases are the principal mechanism by which streams dissipate excess stream power. This may not be the case, however, in sand-bed channels with more resistant banks, where vertical processes may dominate (Simon, 1992).

Maximization of friction factor

Darcy–Weisbach friction factor data derived from channel cross-section surveys and WSPRO modelling were generated for (1) the degrading Elk Rock reach on the debris avalanche of the upper North Fork Toutle River; (2) the aggrading lahar-affected Salmon B reach on the lower North Fork Toutle River; and (3) Burma Gorge, on the Toutle River main stem. In all cases, maximization of friction factor at the Q_1 discharge was not indicated by the data (Figure 12). In fact, the Darcy–Weisbach friction factor (Equation 7) generally decreased slightly with time for these reaches.

The empirical friction factor relation derived for gravel-bed streams (Equation 12) by Bray (1979) was used by Chang (1980) and by Thorne *et al.* (1988) with the minimum stream power hypothesis to predict stable channel geometries. It is used here for the period when gravel-sized particles began to dominate the channel bed (after 1983). Friction factor, as calculated from Bray (1979), displays a non-linear increase for the three reaches, but not one that could be considered as a maximization function since the lines of best fit are concave upwards during the period of record (Figure 12). Clearly, this trend cannot continue indefinitely; it appears that friction factor increases but does not maximize.

The shapes of the friction factor trends are, in part, related to human interference. Sites TL1100 and NF540 are located downstream from the sediment retention structure (SRS) that was closed in November 1987 on the North Fork Toutle River just above the mouth of the Green River (Figure 1). Bed-material samples taken in 1992 at a site 1·5 km downstream from the SRS had the coarsest D_{50} (104 mm) in the basin. Although sites TL1100 and NF540 are considerably further downstream (Table II), the more rapid rate of bed-material coarsening beginning in month 90 at these sites (Figures 6 and 12) can be partly attributed to the closure of the SRS. Still, in all reaches the temporal increase in bed-material particle size, combined with reductions in hydraulic depth, caused 'f', as predicted by Bray's (1979) equation, to increase with time (Figure 12).

The disparity between friction factor values can be attributed to differences in the underlying controls of friction factor in the two equations (Equations 7 and 12). Resistance to flow comprises roughness related to grain size, bedforms, gravel–cobble clusters and, ultimately, pool-riffle morphologies, channel form, and free-surface distortion. Computations of the Darcy–Weisbach friction factor (Equation 7) account for all sources of flow resistance. In contrast, Bray's (1979) relation (Equation 12) calculates a grain-roughness type of friction factor that is an empirical function of relative bed roughness (y/D_{50}) that implicitly includes some average account of other sources of flow resistance. Although it has been generally accepted that bedforms either do not form or are poorly developed in sediments coarser than 2–10 mm, Dinehart (1992b) has shown that dunes can be particularly well developed in the North Fork Toutle River, attaining heights as great as 45 cm in bed sediments with mean D_{50} between 20 and 30 mm during 1989–1990. During this same period, Pitlick (1992) observed gravel bedforms at $\tau/\tau_c > 3$ that reached 20 per cent of the flow depth on the North Fork Toutle River. Dune-like bedforms were also observed in finer gravels by Fahnestock and Bradley (1973), and coarse-gravel bedforms 2·4 m high were reported by Galay (1967) and described as dunes. Using flume data and sediments with D_{50} of 11·5–44·3 mm, Bathurst *et al.* (1987) related the formation and destruction of gravel antidune and alternate bar bedforms to ranges of critical unit water discharge. Willetts (1993) refers to 'gravel clusters' in his flume experiments using 0·1–7·0 mm sediment and attributes temporal increases in flow resistance to first armouring, and then cluster formation and particle rearrangement on the channel bed.

Recognizing the existence of a range of coarse-gravel bedforms and clusters, the underestimates of 'f' using Bray's (1979) function may be due to its failure to account for the shifting dominance of bedform, cluster, and other types of roughness during the adjustment period. Bedform height (amplitude) is controlled by factors such as flow depth, boundary shear stress relative to critical shear stress (excess shear), and grain size. The maximum size of material that can form bedforms, however, is undefined because of a lack of measurements at high flows. Adjustment of channel geometry and hydraulic properties with time in the Toutle River

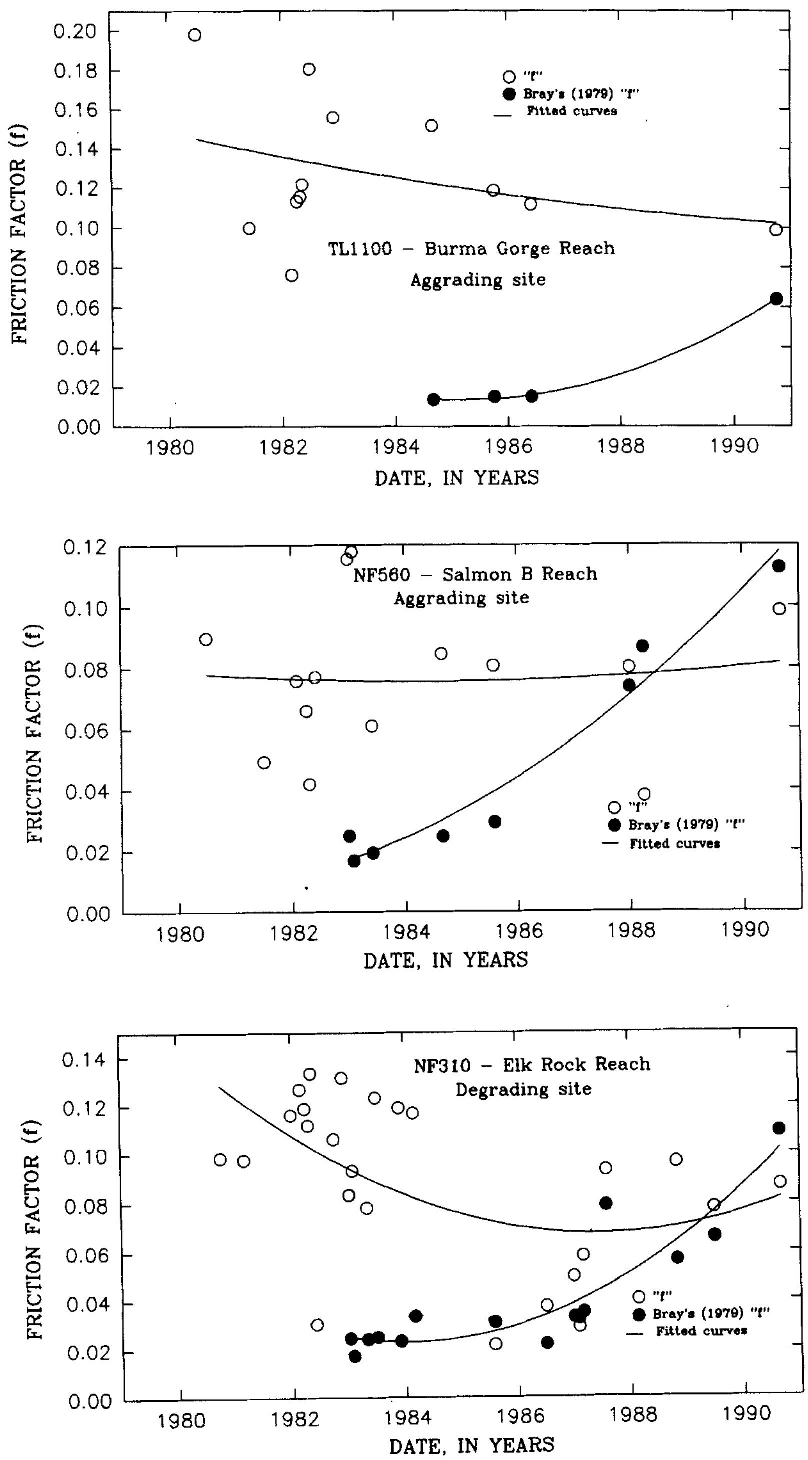


Figure 12. Temporal trends of the Darcy–Weisbach friction factor ('f ' from Equation 7) and from Bray (1979) (Equation 12) at TL1100 (Burma Gorge Reach), NF560 (Salmon B Reach), and NF310 (Elk Rock Reach). Note the lack of a maximization trend

system resulted in decreases in flow depth and excess shear stress, coupled with increases in grain size, so that the potential for bedform development at the Q_1 discharge probably became small in the study reaches by 1991–1992. By 1991–1992 τ/τ_{cr} was generally less than 2·0, below the value identified by Dinehart (1992b) for the development of coarse-gravel bedforms (Figure 5). Grain roughness increased markedly with time (Figure 12) and by 1991–1992 was expected to be the dominant source of flow resistance. Thus, the end points of Bray's (1979) 'f ' should not require upward adjustment to account for this additional resistance of bedforms, and should approach the actual Darcy–Weisbach 'f '.

The maximization of friction factor, as defined by Davies and Sutherland (1983) (Equation 7), does not apply for adjustment of the unstable reaches modelled in the Toutle River system (Figure 12). In fact, derivation of the maximum friction factor hypothesis by Griffiths (1984) shows that the friction factor does not display a numerical maximum when channel depth, width and slope are dependent variables. Bettess and White (1987) similarly found no maximum of friction factor in their analysis of regime relations.

Extremal hypotheses and prediction of stable channel geometries

There is poor agreement between observed and predicted 1992 channel geometries using Chang's (1980) minimum stream power approach (Equations 17, 18). Only at the most upstream site (NF120 in the Coldwater reach) is the predicted width within 50 per cent of the observed value. In general, the channel is much wider than predicted, with the result that width is underpredicted by as much as 90 per cent. Width is significantly overpredicted at only one location (NF130, Coldwater reach; kilometre 76·8). Depths are generally overpredicted, except in the uppermost Coldwater reach where the observed depth is seriously underpredicted. As an independent check on Chang's (1980) minimum stream power method, expected stable-channel widths and depths were also calculated using Hey and Thorne's (1986) empirical equations for gravel-bed rivers with mobile bed materials. These indicated a regime width of 57 m and a depth just over 2 m for the North Fork Toutle River. Predicted regime width and depth for the Toutle River main stem were 73 m and 2·4 m, respectively. These predictions are broadly comparable to those of the minimum stream power approach. Predicted values indicate stable geometries that would be narrower than predisturbed channels, which had width-to-depth ratios ranging from 60 to 100 (Collings and Hill, 1973). On the basis of these results, it must be concluded that: (1) these regime-type approaches are not well suited for the Toutle River system's current (1992) conditions, which are characterized by extremely wide, braided streams; and/or (2) the Toutle system had not, by 1992, achieved a stable geometry.

Potential sources of predictive errors

The selection of the Q_1 discharge as a representative, high in-bank flow may be questioned. The Q_1 does not equate directly to bankfull flows as stipulated in the minimum-stream-power approach. In fact, the stage height and magnitude of the bankful discharge changes with time as the channel undergoes adjustment, making it impossible to define a time-invariant value. In practice, the definition of a bankfull discharge is essentially arbitrary in braided reaches because of the complex and constantly changing morphology of the channel cross-sections. It is for this reason, however, that a relatively high 'channel-forming' flow (Q_1) was selected.

Empirical data on the temporal trends of the rate of energy dissipation (Figure 10) and the Darcy–Weisbach friction factor (Figure 12) permit identification of a potential source of the predictive error. Chang's (1980) minimum-stream-power approach relies on three criteria: (1) the concept of minimum stream power; (2) Bray's (1979) resistance function (Equation 12); and (3) sediment transport equations (Equations 13–15). Data from the Toutle River system indicate that the first criterion, minimization of stream power expressed as minimization of the rate of energy dissipation (Bettess and White, 1987), is applicable. Values of the Darcy–Weisbach friction factor, estimated using Bray's (1979) resistance equation, approach values observed during 1991–1992 (Figure 12) and thereby satisfy the second criterion. The problem, therefore, seems to lie in the third criterion, the selection of the sediment transport equation and the prescribed sediment inflow. This conclusion was also reached by Bettess and White (1987).

The predictive abilities of the selected sediment transport function for a given fluvial system are paramount because, in the minimum-stream power-approach, slope is predicted by rearrangement of the

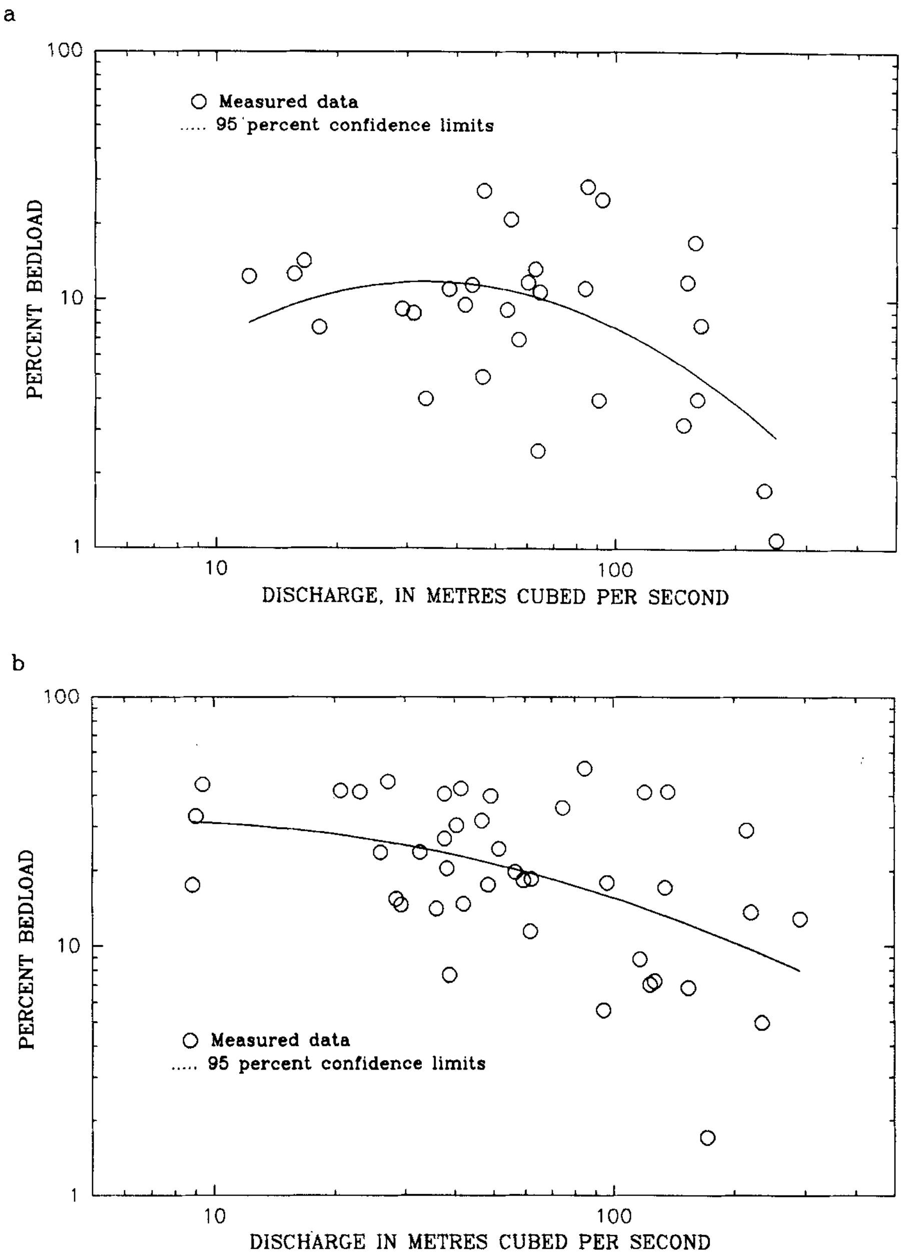


Figure 13. Percentage contribution of bed load to total load for (a) the Toutle River main stem, and (b) the North Fork Toutle River

sediment transport equation to express slope as a function of the bed-material characteristics and the prescribed upstream inflow of bed load or bed-material load. With flow discharge held constant, the prediction of slope equates to prediction of stream power. This is the basis for the predictive multiple-regression regime-style equations (Equations 17 and 18) derived by Chang (1980).

The Parker (1978, 1979) equations used by Chang (1980), Griffiths (1984) and Thorne *et al.* (1988) are based on the assumption that in gravel-bed streams, the sediment load is transported primarily as bed load. However, bed load accounts for only a fraction of the total load for a given discharge at sites on the North Fork Toutle River and the Toutle River main stem (Figure 13). In 1992, total sediment discharge

from the Toutle River system was about 8·5 million tonnes (Table III), indicating that significant amounts of geomorphic work were taking place that would be unaccounted for by a bed-load equation. At the Kid Valley and Tower Road gauges, the median percentage of sand in suspended-sediment samples was about 50 and 60 per cent, respectively. A bed-material discharge equation that is applicable to both sand- and gravel-sized sediment, such as Ackers and White (1973), might therefore be more appropriate for the Toutle River system.

Bettess and White (1987) demonstrated that in combination with the minimum-stream-power approach, the Ackers and White (1973) equation gave results comparable to conventional regime relations, whereas the Parker (1978, 1979) equations did not. However, direct application of these findings to the Toutle River system is problematical because of the use of low inflow concentrations of suspended sand (10 ppm) by Bettes and White (1987). Sediment-transport data indicate that between 1982 and the closure of the SRS in November 1987, concentrations of suspended sand at the 50 per cent exceedance probability were at least 1000 ppm. (R. Dinehart, U.S. Geological Survey, written communication, 1993).

With much of the geomorphic work during 1991–1992 involving channel widening and floodplain formation by mechanical failure and retreat of stream-side hummocks (20–40 m high), at least 50 per cent of the material delivered to channel beds and terrace surfaces is sand. This sediment, which is finer-grained than the mobile gravel and cobble pavement of the channel bed, is available for entrainment as suspended load by flows that spread across low-terrace surfaces. Because of heightened concentrations of sand, sediment transport capacity and upstream sediment inflow can be higher than expected for a given flow discharge (Ferguson *et al.*, 1989). Comparative tests using combinations of different sediment transport and resistance algorithms are beyond the scope of this paper but would constitute a valuable contribution to understanding the validity and applicability of extremal hypotheses in predicting stable channel geometry. Work by Bettess and White (1987) represents an important step in this direction.

The active channels of the Toutle River system appear to be approaching equilibrium; however, it seems that the final stage of channel evolution will not be complete until a definable floodplain has formed. It is expected that this will occur by reincision and consequent channel narrowing as sediment loads decrease from the upper debris avalanche and by continued retreat of stream-side hummocks.

DISCUSSION

Results from the Toutle River system highlight the fact that adjustments can temporarily produce configurations that bear little or no relation to those of a regime geometry. Such conditions may constitute a temporary equilibrium which is apt to be disturbed as the channel continues to evolve. New phases of adjustment may be characterized by renewed incision through breaching of bed armour, or the influx of significant volumes of sand-sized sediment from upstream erosion. In fact, long-term stability cannot be assured until floodplain formation and revegetation of riparian zones re-establish effective controls on the destabilizing effects of large events. The evidence from morphological features produced by the 1857 eruption of Mount St. Helens is that revegetation may take centuries rather than decades. The processes of floodplain formation by failure and retreat of hummocks on the debris avalanche deposit and the lahar deposits in the lower reaches will continue to supply copious amounts of sediment to the system for many years. The maintenance of a relatively high sediment load has delayed reincision and the consequent reduction in bankfull channel widths in many upstream reaches. Under these circumstances, it is not surprising that predictions based on stable conditions bear little relation to the exceptionally wide and shallow channels of the Toutle River system.

The results of this aspect of the study demonstrate the inability of the regime-style approach to predict channel geometry even when (1) applied and critical boundary shear stresses have converged to a considerable degree, and (2) unit stream power and the rate of energy dissipation have approached minimum values. It may be concluded that the only possible way to develop a means to account for time-dependent adjustments is with an approach which physically accounts for the dynamics of channel evolution from an unstable to a stable state. Further understanding of the links between adjustment processes and channel morphology require advances in two areas. Firstly, an improved ability is required to characterize and model the

processes of selective entrainment and downstream sediment sorting that are responsible for the coarsening of the bed and for forming and destroying bedforms and pebble clusters. Secondly, recognition of the important roles played by processes of bank erosion, bank collapse and lateral sediment diffusion, which are responsible for width adjustment and associated plantform changes, highlights the need to be able to describe and model these processes effectively. Development of numerical models that account explicitly for bed armouring, cluster formation, widening through bank erosion and collapse, and coupling of mass-wasting processes with in-channel sediment-transport dynamics represents the best chance for improved explanation and prediction of channel geometry in an unstable fluvial system like that studied here.

SUMMARY AND CONCLUSIONS

Disruption of the Toutle River system by the catastrophic eruption of Mount St. Helens on 18 May 1980, resulted in large-scale channel instabilities and marked channel adjustments. Changes in channel geometry were monitored by repetitive cross-section surveys that were used as input data to investigate temporal adjustments in flow energy and hydraulic properties. Minimization of unit stream power and the rate of energy dissipation were applicable to dynamic adjustment in the study reaches, with channel widening being the dominant process. Widening reduced hydraulic depths and flow velocities through increases in the size of the wetted perimeter resulting in increased flow resistance. These effects were coupled with bed-material coarsening, which is an important variable in channel adjustment. Bed-material coarsening and the formation of a mobile pavement have two principal effects: (1) it causes critical shear stress to increase with time; and (2) it results in preferential erosion of the non-cohesive channel banks which become finer-grained relative to the channel bed. The non-linear increase in critical shear stress with time is shown to approach a balance with boundary shear stress at the Q_1 discharge, and represents the approaching of an equilibrium condition.

Attempts to use extremal hypotheses to predict stable channel geometries in the Toutle River system were unsuccessful. Maximization of the Darcy–Weisbach friction factor did not occur in the Toutle River system, although relative bed roughness did increase with time. In comparison to 1991–1992 channel geometries, the minimum stream power approach generally underpredicted channel widths and overpredicted channel depths. This was attributed to two causes: firstly, that reaches of the Toutle River system are not in regime and a dynamic equilibrium will not be attained until a new floodplain has been formed; and secondly, that the ability of the sediment-transport equations to predict energy slope accurately is limited by their failure to account for the transport of high concentrations of sand (emanating from failed channel banks) as bed-material load and, therefore, they underestimate upstream sediment inflows. In the minimum-stream power approach, the prediction of energy slope equates to the prediction of stream power because discharge is a constant. Consequently, the coupling of bank-failure processes and sediment supply with bed-material transport should be incorporated explicitly into the sediment-load algorithms. This is where advances in predictive capabilities and the application of extremal hypotheses could be improved.

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